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Dynamic Topography Change of the Eastern U. S. since 4 Ma: Implications for Sea Level and Stratigraphic Architecture of Passive Margins

David B. Rowley[1], Alessandro M. Forte[2], Robert R. Moucha[2,3], Jerry X. Mitrovica[4], Nathan A. Simmons[5] and Stephen P. Grand[6]

[1] Department of the Geophysical Sciences
5734 S. Ellis Ave.
The University of Chicago
Chicago, IL 60637

[2] GEOTOP - Université du Québec à Montréal
CP 8888, succursale Centre-Ville
Montréal, Québec
Canada H3C 3P8

[3] Department of Earth Sciences
204 Heroy Geology Laboratory
Syracuse University
Syracuse, New York 13244, USA

[4] Department of Earth and Planetary Sciences
Harvard University
20 Oxford Street
Cambridge, Massachusetts 02138, USA

[5] Atmospheric, Earth & Energy Division
Lawrence Livermore National Laboratory
Livermore, California 94551, USA

[6] Jackson School of Geological Sciences
University of Texas at Austin
Austin, Texas 78712, USA

Abstract

The mid-Pliocene stratigraphy from Virginia south through Florida record a marine flooding surface associated with transgression as far west as the wave-cut Orangeburg, Chippenham and Thornburg scarps. These wave-cut scarps which at the time of deposition would have been horizontal, and the associated flooding surface that would have been gently sloping towards the east are now draped over a warped surface with a maximum amplitude of 55 to 60 m or more. Models of dynamic topography using mantle convection simulations predict the amplitude of distortion for this flooding surface. Together with other, but distintly smaller effects associated with glacial isostatic accommodation, and flexure related to offshore sediment loading and onshore erosional unloading, account for the deformation of this surface within the past 3 million years. We find that in the last several million years, dynamic topography has dominated over other traditionally considered mechanisms accounting for coastal plain and proximal shelf stratigraphic architecture, confounding attempts to use local or even regional stratigraphic relations as references for long-term sea level determinations. As a consequence, inferences of Pliocene global sea level heights, in general, and stability of Antarctic ice sheets, in particular, cannot be deciphered from coastal plain data in the absence of an appropriate mantle dynamic reference frame.

Introduction

The continental margin of the east coast of the U.S. is the archetypal Atlantic-type or passive-type continental margin[1]. Passive-type margins are assumed to be underlain by mantle that is entirely passive, contributing nothing beyond thermal boundary conditions associated with thinning and stretching that drives the long-term thermal subsidence of the margin and background heat flux controlling exponential rather than half-space cooling[2]. The east coast margin is generally interpreted as being characterized by a simple stratigraphic history with sedimentary packages increasing in thickness away from a zero edge near the coast, the architecture of which is assumed to be virtually entirely controlled by the interplay between exponentially decreasing, thermally-driven, subsidence together with sediment loading, compaction, and sea level variations of varying amplitude and frequency as a function of time, modulated by variations in sediment supply[3,4]. Flexural responses of the lithosphere due to off-shore sediment loading[5,6] and less frequently onshore erosional unloading[7] are also recognized as potentially important contributors to the stratigraphic architecture of these types of margins [3,5,6]. These assumptions underpin the rationale for the use of this margin, in particular, in determining presumed global longer-term (≥ 0.1 Ma) sea level variations [3,5,6,8-10].

In contrast to this view, Moucha et al.[11,12] emphasize that the mantle cannot be assumed to be a passive player, and explicitly argue that mantle-derived dynamic topographic contributions that vary spatially and temporally, in amplitude and sign, contaminates the geological record in ways that make it impossible to uniquely invert for the global long-term sea level signal from these and other settings. The potential effects associated with mantle-derived dynamic topography and its variations in time have also been discussed by others [13-15]. These other models[13-15] are in broad agreement and are everywhere characterized only by retrodicted dynamic topographic subsidence over the last 50 plus million years. Based on this Müller et al. [14] concluded that Miller et al.[10] substantially underestimated long-term sea level variations by ignoring the dynamic topography change contribution to the evolution of the New Jersey margin. The Moucha et al.[11,12] results differ from those cited above because their analysis incorporates contributions from both flow associated with the negative buoyancy of the subducted Farallon slab, assumed by others[13-15] to be the only source of dynamic topography-related flow, but also the coupled shallower westward flow of hotter mantle. This hotter mantle contributes, at least locally, significant positive vertical components to the dynamic topography that dominate over the Farallon slab-related dynamic topographic subsidence along the east coast (Figure 1, see also Supplementary Information).

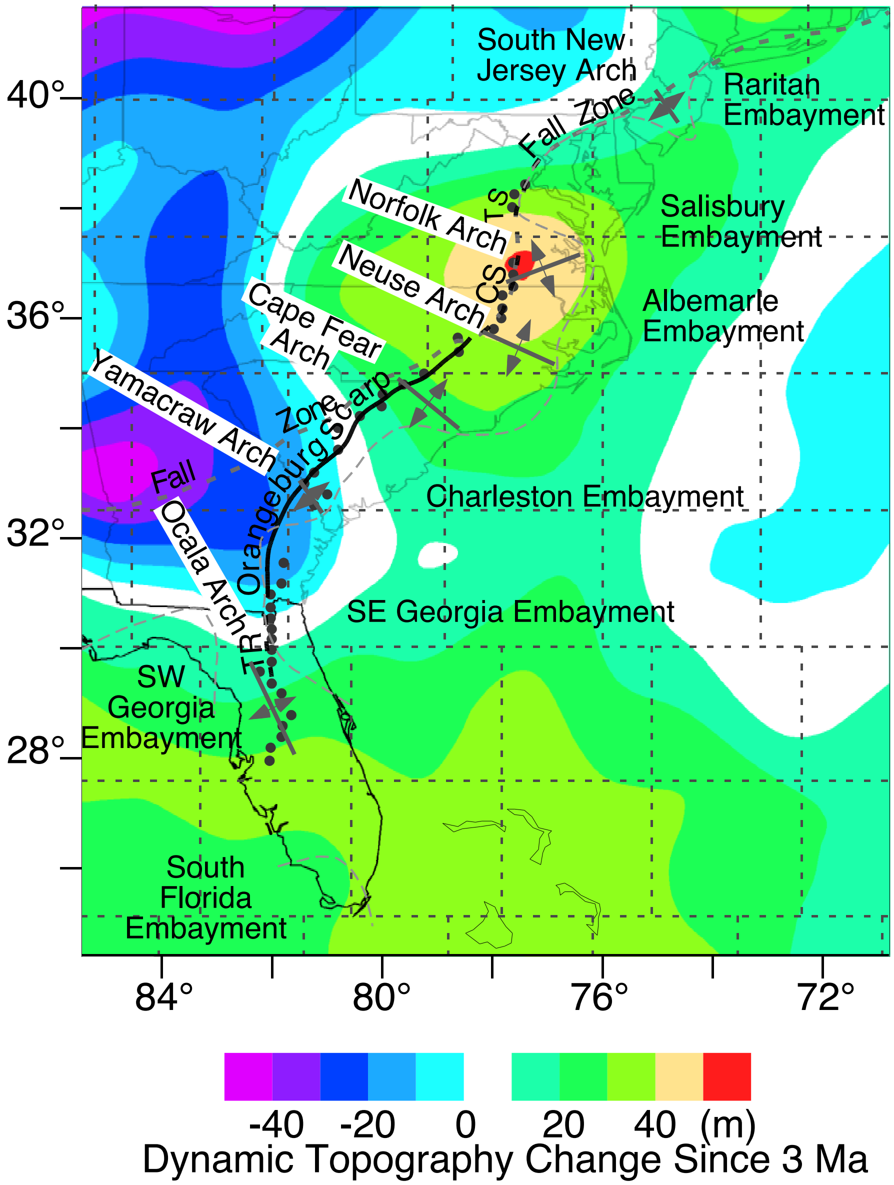

Figure 1. Locations of features associated with east coast Coastal Plain geology after [16]. CS is the Chippenham scarp, and TS is the Thornburg scarp. The Fall Zone marks the approximate landward erosional edge of the Early Cretaceous to Cenozoic Coastal Plain strata. Locations (gray dots) with preserved mid-Pliocene (Yorktown, Duplin, Chorlton, and Cypresshead Formations) strata. Color image is the distribution of retrodicted dynamic topography change

since 3 Ma based on model TX2007 and the V2 vicosity profile. For reference to Figure 2, this model retrodicts the least amplitude of change for regions north of 30°N. Dashed gray rectangular boxes outline the underlying resolution of the Simmons et al. [43,44] jointly inverted seismic tomography.

We test the predictions of the Moucha et al. [12] modeling against others[13-15], by examining the current elevations of the mid-Pliocene shoreline and associated marine sediments that reflect a major flooding event preserved along the east coast of the United States. Coastal Plains geologists have long recognized various lines of evidence for differential vertical motions along the length of the Coastal Plains from Florida to New Jersey [17-20] (Figure 1). These studies[17-20], among many others, have highlighted the considerable geological evidence supporting significant (10's to 100's m) differential vertical motions even in relatively recent times. Thus, for example, Winker and Howard [17] traced the Pliocene Trail Ridge-Orangeburg scarp (Figure 1) from northern Florida to North Carolina and showed that this nominal Pliocene paleoshoreline-related wave cut scarp, at least north of Florida, varies by more than 60 m in elevation with varying lesser amounts of distortion for younger Pliocene and Pleistocene shorelines. We suggest below that this deformation is a manifestation of the interplay between mantle-related changes in dynamic topography as a primary contributor, together with other processes typical of "passive-type" margin evolution.

Dynamic topography-related deformation of the Orangeburg Scarp also has direct implications for estimates of Pliocene maximum sea level height. Dowsett and Cronin[8] estimate a Pliocene sea level height of 35±18 m above current sea level, based on their reconstruction of the amplitude of post-Pliocene deformation of the Orangeburg Scarp in North and South Carolina. This [8] estimate has engendered concern about the stability of the East Antarctic ice sheet (see commentary by [21]).

Coastal Plain Geology

The Coastal Plain of the east coast of the U.S. is characterized by a sequence of marine and non-marine sedimentary units that range from at least Early Cretaceous to Present in age. These units generally thicken eastward beneath the Atlantic shelf where in the Baltimore Canyon Trough and Carolina Trough total post-Paleozoic sediment thicknesses can exceed 12 km [22-25]. This package of sedimentary units unconformably overlies various pre-Mesozoic crystalline rocks, as well as Triassic/Jurassic rift-basin strata, and pinches out to the west along the Fall Zone (Figure 1). Significant differences in depositional sequences along strike across arches and within adjacent basins [16,26], referred to by coastal plains geologists as embayments (Figure 1), are typically ignored in modeling studies, as is evidence that embayments and arches have shifted as a function of time [27]. Detailed models of the depositional architecture of various segments of this margin have been developed based on combinations of seismic stratigraphy and drilling [10,24,25] (among many others) in order to better understand its evolution [3,5,6,28]. One motivation for these efforts has been the attempt to determine global long-term sea level history by solving for the contributions of thermal subsidence, sediment-loading and compaction, flexural loading (but not unloading), and sediment delivery and by assuming that the only remaining unknown term is the contribution from changes in sea level with its attendant water-loading contribution to the overall subsidence. These studies specifically focus on the New Jersey segment of this margin [3,5,6,9,10] see also [29] focusing on west Africa). Possible contributions from mantle-flow induced dynamic topography have been largely ignored.

Dowsett and Cronin [8] and Cronin [18] among others, have emphasized that the mid-Pliocene Orangeburg Scarp may provide a marker for the determination of the height of sea level at this time, with potential implications for the extent of melting of the East Antarctic Ice Sheet. Specifically Dowsett and Cronin [8] have argued that after correction for ~50±18m of uplift extrapolated based on empirical estimates of stream incision rate[30], the Orangeburg Scarp would have been at an elevation of 35±18m which they infer to have been the height of the mid-Pliocene sea level. Given that best estimates for the global eustatic sea level change associated with complete melting of Greenland and West Antarctic ice sheets is only +14m [31], this height would imply potentially considerable (~1/3) melting of the East Antarctic ice sheet in the mid-Pliocene and hence much less stability of East Antarctica than is often inferred [21].

Select localities of the mid-Pliocene strata are shown on Figure 1 based on outcrop and shallow bore holes along the length of the Orangeburg Scarp, and its northern continuations, the Chippenham and Thornburg Scarps in Virginia. These localities represent the westernmost extent of these units at each latitude and thus provide independent spatial control on the distribution of shallow marine sediments adjacent to these scarps. The elevations of these localities as a function of latitude are plotted in Figure 2. Where these units are in the subsurface a simple Airy correction for loading by overlying sediments has been applied that averages approximately a +5m adjustment to the borehole elevations. What is quite obvious, and in complete agreement with Winker and Howard[17], is that Pliocene strandline and or adjacent shallow shelf sediments are not preserved at constant elevation along the Coastal Plain. This clearly demonstrates that the Orangeburg Scarp is not a good reference for sea level determinations of the Pliocene, but may well be an ideal marker for learning about other processes that may play roles in warping this sea level marker since its creation between 4 and 3 million years ago [20].

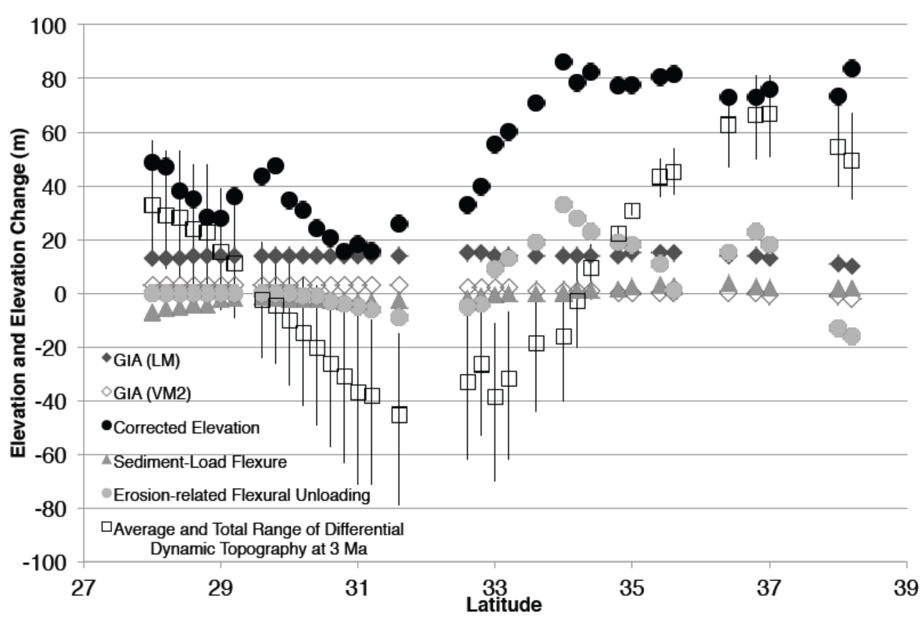

Figure 2. Current elevations and load corrected elevations of shallow boreholes (filled black circles) and corresponding amplitudes of elevation change since 3 Ma for localities shown in Figure 1 along or proximal to the Orangeburg, Chippenham, and Thornburg Scarps as a function

of latitude. Two estimates of topographic change associated with GIA are shown based on results from Raymo et al[31]. Differential dynamic topography change is shown as the average of our 4 models with vertical bars representing the total range represented by the models.

Modeling East Coast Topography since the Miocene

Several processes have been identified that may influence the post-depositional elevations of paleo-shorelines and shallow marine sediments across the Coastal Plain. These include remaining post-glacial isostatic adjustments[18,31], flexural warping due to offshore sediment loading and correlative erosion-related flexural unloading[7,32], karstification-related flexural unloading[33], and dynamic topography change[12-15]. Given that carbonates represent a fairly limited portion of the stratigraphy in regions north of Florida, and therefore any karstification effects are of limited significance for regions not in Florida, we will ignore this effect in the following discussion. We will treat the remaining three in succession.

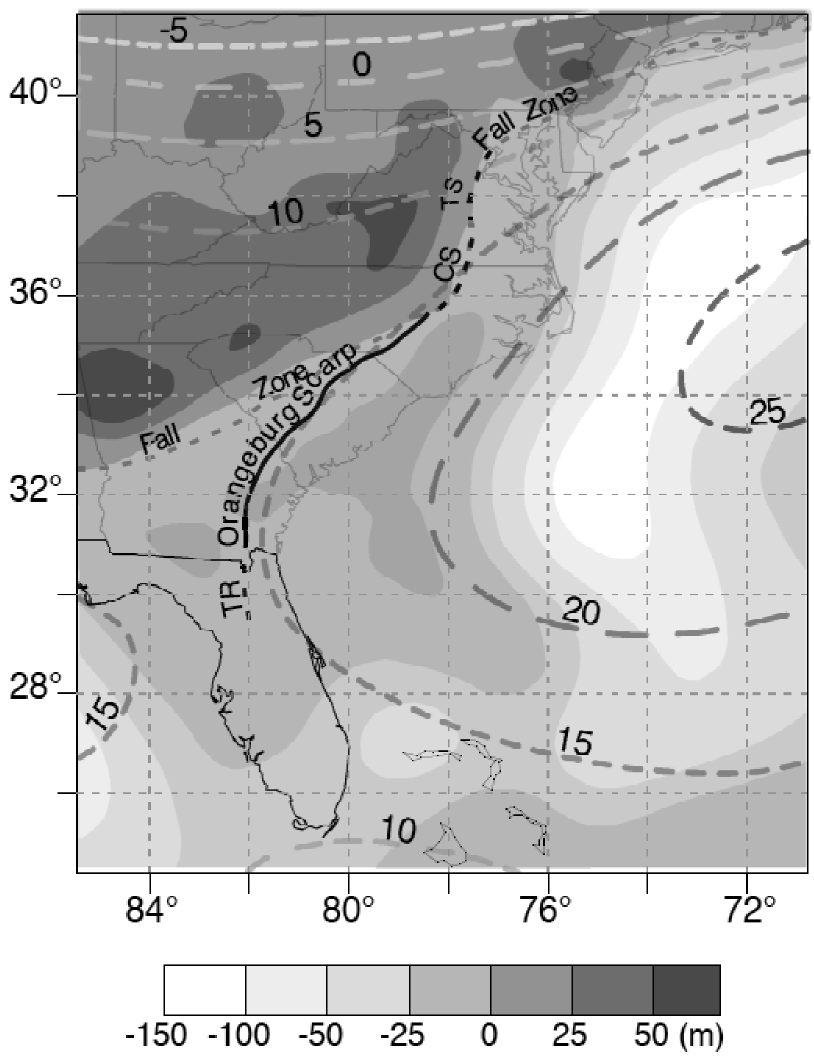

Figure 3. Estimates of the amplitude of the remaining glacial isostatic adjustment and combined effects of flexural loading and unloading. GIA estimate based on the LM solution of Raymo et al. ([31], See their Figure 3c) shown by labeled dashed contours. Note that the 15m contour essentially coincides with the Orangeburg Scarp. Shading represents an estimate of the

amplitudes of flexure due to post-3Ma sedimentation in offshore basins and unloading due to post-3Ma erosion. Subtraction of these amplitude from the modern topography yields an estimate of the non-glacially and non-flexurally influenced topography of this region at 3 Ma.

Glacial Isostatic Accommodation

Post-glacial isostatic adjustment (GIA) refers to the ongoing adjustment of the Earth as a consequence of the Late Pleistocene glacial cycles. The gravitational, deformational, and rotational aspects of this adjustment are manifest in global-scale sea level variations, which include the current apparent sea-level fall (or post-glacial uplift) of previously glaciated regions, and apparent sea-level rise (land subsidence) of the peripheral bulges that encircle these zones of ancient ice cover. The U.S. east coast is largely located on the peripheral bulge of the Laurentide ice complex, and it is subject to crustal subsidence, and apparent sea level rise, that has continued unabated since the end of the last glacial maximum at 21 ka[31]. Raymo et al. [31] computed the amplitude of this remaining glacial isostatic adjustment along the U.S. East coast (Figure 3). Upto 15 m of the current topography along the Orangeburg Scarp could be associated with GIA with the amplitude decreasing westward and increasing offshore. Subtraction of this GIA component from the modern topography thus yields an estimate of the long term non-glacial isostatic topography of this region. Note that the 15 m contour parallels the Orangeburg Scarp over the majority of its length, and thus although GIA impacts its current elevation it does not account for any significant amount of the vertical distortion along it. Uncertainty in the viscosity of the mantle and Laurentide ice history leads to uncertainty in the estimate of the remaining GIA along the East coast of the U.S. [31]. A second mantle viscosity profile (VM2; [34]) also considered by [31] (see their Figure 2c) predicts essentially no change in elevation due to GIA along the Orangeburg Scarp (Figure 2). We use the values derived from the LM model for purposes of our calculations below.

Flexural Loading and Unloading

There is a substantial volume of sediment in offshore basins of various ages along the Atlantic and Gulf coasts of North America [22-24]. Much of this sediment is either Triassic/Jurassic or Middle Miocene and younger in age [22,24,35]. Lithospheric flexure associated with this loading, and most specifically with the Neogene, would be expected to have warped the Coastal Plain surface in regions away from the load, resulting in uplift of the surface, potentially including Pliocene and younger shore lines [7]. Isopach maps of the Pliocene and younger sequences are not available for the entire East and Gulf coasts basins, but do exist for the entire Neogene [22] which we use to compute the flexural response to Neogene sediment loading.

We modulate the total Neogene flexural loading by ascribing ~30% of total loading to effects associated with Pliocene (at 5.0 Ma) and younger sediment loads, based on data from the Baltimore Canyon Trough where better stratigraphic resolution allows the distinction between Miocene and post-Miocene sequences [24,35]. We assume a constant rate of sediment-related loading since the Miocene resulting in a 6%/Ma change in the total sediment load since 5 Ma, in accord with available data [35]. The amplitude of the flexural response due to offshore sediment loading along the Orangeburg Scarp is shown in Figure 2 representing our estimate of the total flexural response over the past 3 Ma (see Supplementary Information). Sediment-load related flexure is estimated to effect the topography variably by about 12m along strike, ranging from about 7m subsidence in the south to about 5m uplift in the north (Figure 2).

An additional flexural response related to erosional unloading (see [7]) needs to be included to fully assess elevation changes associated with flexure. However, the amplitude and distribution of erosion in the past 5 Ma is not well known over the region of Figure 1. Local estimates based on catchment wide averages derived from cosmogenic nuclides[36-38] typically average about 20 m/Ma, but range from 0.6 to 57 m/Ma. Local estimates of erosion rate derived from time-temperature history modeling of apatite fission track and apatite U/Th-He dating from around the Appalachians, Blue Ridge and Piedmont yield estimates of around 40±20 m/Ma assuming a regional consistent 25°C/km geothermal gradient over the past ~20 Ma (Rowley and Komacek, unpublished analyses of published AFT and AHe data). For the current analysis we use a slightly modified version of post-50Ma estimate of erosion rates[39]. We suspect that these are maximum estimates of erosion rate[39] based on comparisons of recent fission track inversions of time-temperature histories relative to those used to obtain these estimates[39]. There is considerable uncertainty in the estimate of the spatial pattern of erosion mainly reflecting the limited number of locations where any control exists and uncertainties at those locations associated with the techniques used to derive the local estimates. Combined these lead to considerable uncertainty in our estimate of flexural unloading. Based on current best estimates there is some along-strike variation in the amplitude of erosion-related flexural unloading from +33m to -16m of the Orangeburg Scarp (Figure 2).

The combined effects of flexural loading and unloading are portrayed in Figure 3.

Mantle Flow-related Dynamic Topography

Calculations of the mantle convective flow used in this analysis follow the approach of Moucha et al. [12,40] and Forte et al. [41,42], in which seismic tomography models (TX-2007 from [43] and TX2008 from [44] together with a range of geophysical observables (present-day surface topography, free air gravity, plate velocities, and core-mantle boundary excess ellipticity) have been jointly inverted to yield a geophysically consistent estimate of the 3-D distribution of density in the mantle [44]. For the current exercise we employ two different models of the radial distribution of viscosity, V1 and V2, (see [12], that together with the two different inversions of seismic velocity to density provides four alternative models of the dynamic topography as a function of time.

The 3D distribution of buoyancy, which when integrated with estimates of the radial distribution of viscosity, allows the computation of the instantaneous flow field. As discussed in more detail in Moucha et al. [12,40] we then iteratively compute a backward advection solution brought forward in time with a full convection calculation to estimate the evolution of dynamic topography. Dynamic topography reflects the contribution of vertical stresses arising from flow in the mantle acting on the base of the lithosphere that result in non-isostatic variations in topography. The difference between the present dynamic topography and estimates of past dynamic topography result in retrodictions of the change in dynamic topography as a function of time (Figure 1). Figure 2 portrays the full range of retrodictions of differential dynamic topography over the past 3 Ma based on the four models discussed above (see [40] for a discussion of these) at each locality.

The current load-corrected height of the Orangeburg scarp-related sediments and dynamic topography change since 3 Ma show a good correlation in their latitudinal variations. Both dynamic topography change and current height are high in the south, decrease towards the north in the Southeast Georgia Embayment (~31°N), and then rise again farther north over the Cape Fear, Neuse and Norfolk arches (Figure 1). The height of the Orangeburg Scarp rises more quickly starting at about 32°N than does the much smoother dynamic topography change since 3 Ma. As is clear from Figure 2, the differential dynamic topography has the largest amplitude signal of all of the effects considered, and hence contributes the majority of the correction to the retrodicted elevations of the Orangeburg Scarp.

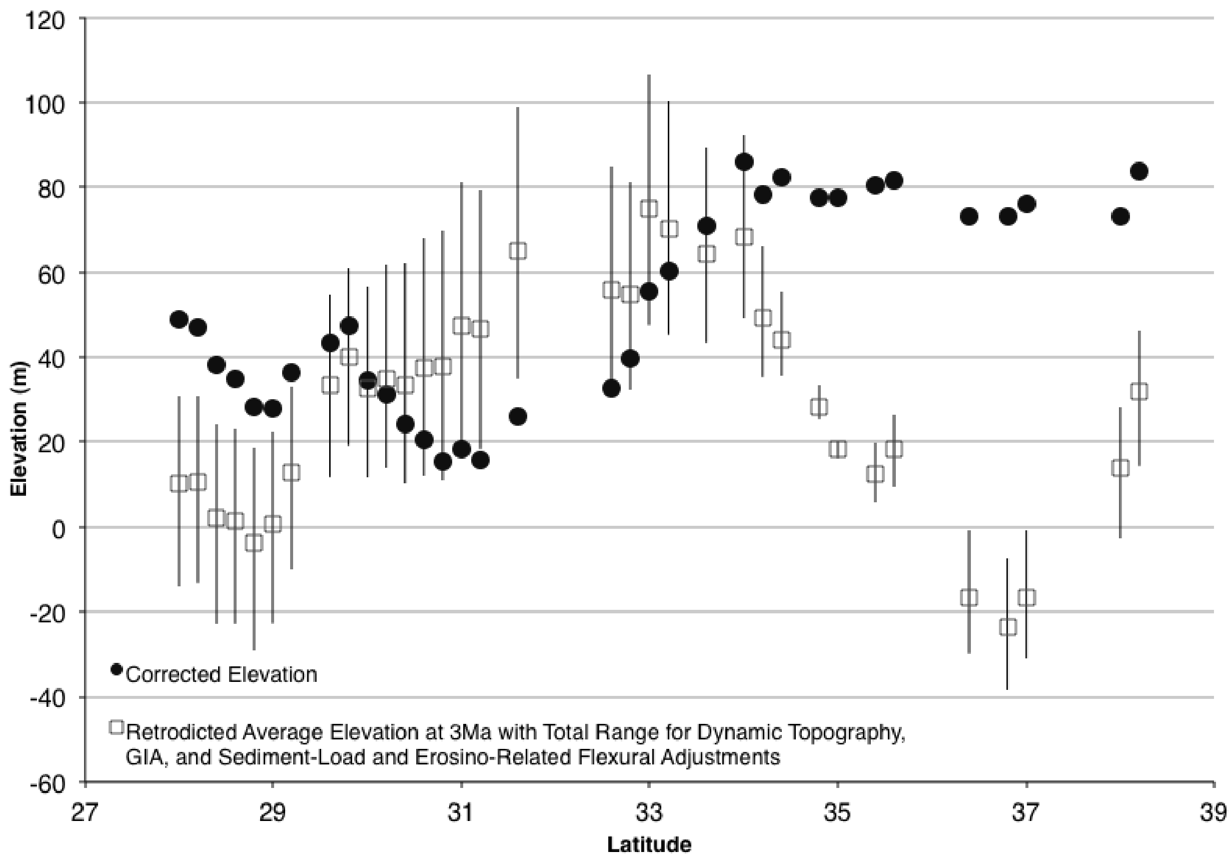

Figure 4. Elevations of the westernmost localities of Yorktown, Duplin, Reysor, and Cypresshead Formations proximal to the Orangeburg and related scarps as a function of latitude. Black dots are present day elevations of outcrop exposures and Airy-load corrected elevations of shallow boreholes. Vertical lines represent the full range of the GIA + Flexural Loading and Unloading + Dynamic Topography corrected elevations at 3 Ma.

Retrodicted Paleogeography of the late Early Pliocene

Figure 4 is a remarkably stringent test of the mantle flow retrodictions of dynamic topography change in that the horizontal spatial scale of the geological features is known to within a few kilometers or less, whereas the underlying joint seismic-geodynamically constrained tomography has a horizontal spatial resolution of 250 km by 250 km[44] and thus averages at a length scale much greater than the geological data being considered. To place these results in a broader context, Figure 5 shows a larger scale map of the eastern U.S. in which the various contributions from GIA, flexure, and dynamic topography are subtracted from the present day topography to retrodict the paleogeography at 3Ma. The figure demonstrates that on this regional scale there is a good correspondence between geological data that constrain the known distribution of marine mid-Pliocene sediments, inferred shoreline positions based on the geology as interpreted by Krantz [20] and Ward et al. [16], and position of the retrodicted shoreline (0m elevation contour) assuming no change in global eustatic sea level relative to today.

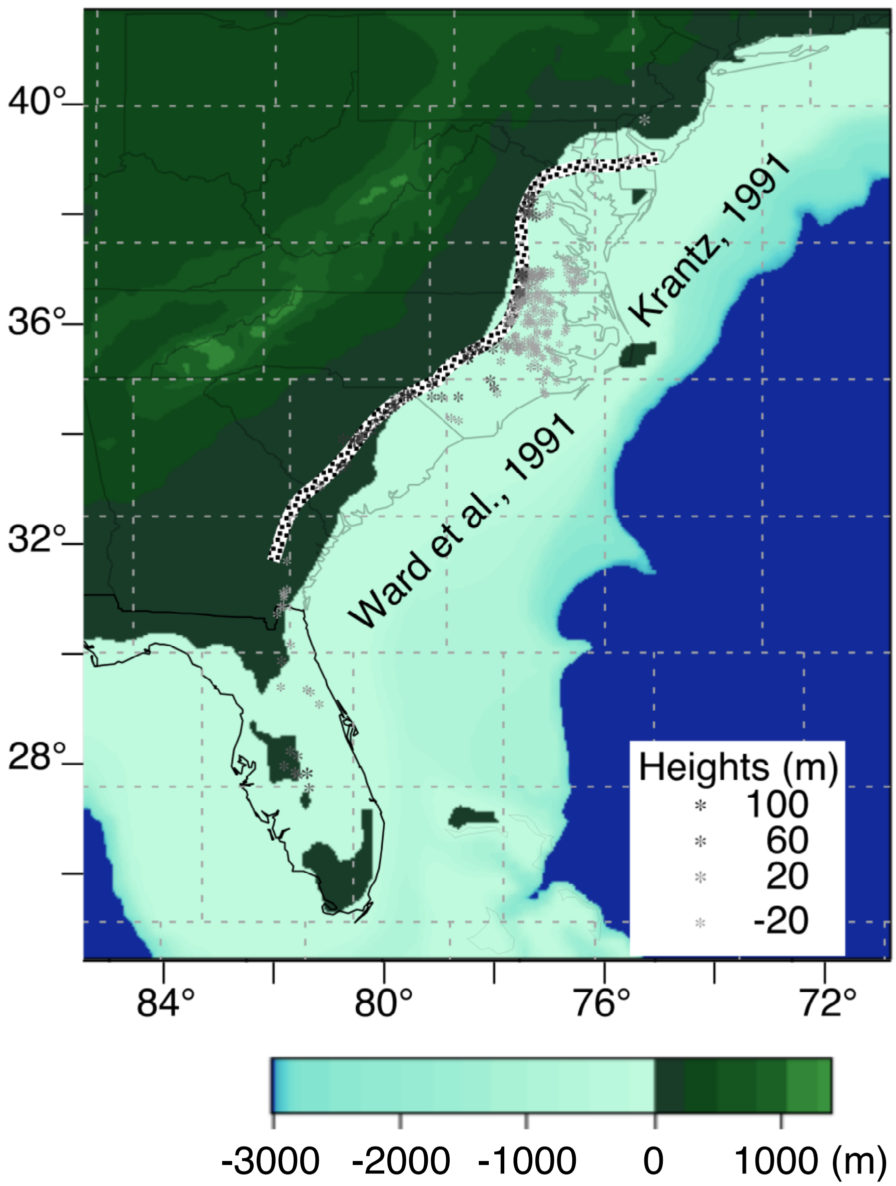

Figure 5. Retrodicted paleogeographic reconstruction of the eastern U.S. at 3 Ma. Retrodicted elevations based on ETOPO1 elevations regridded at 0.2°x 0.2° resolution from which has been subtracted contributions from GIA, flexure at 3Ma, and differential dynamic topography based on TX2007 and V2. Coarse dotted line is the shoreline inferred geologically by Krantz [20] and Ward et al. [16]. Gray asterisks are locations for which there are independent outcrop or borehole constraints on the presence of Pliocene marine sediments. Dashed gray rectangular boxes outline the underlying resolution of the Simmons et al. [43,44] jointly inverted seismic tomography. Asterisks in southern Delaware and New Jersey, respectively, are locations of Pliocene estuarine sediments [45,46].

Discussion and Conclusions

There is increasing interest in and application of models of dynamic topography to a wide range of geological problems. Long-wavelength dynamic topographic subsidence in back-arc regions associated with subducting slabs [47,48], regional topographic uplift associated with ascending mantle "plumes" (e.g.[40,42,49]), and potential implications of dynamic topography for estimates of global sea level determinations derived from specific locations such as the New Jersey margin (e.g. [12-15]) are increasingly being discussed and debated.

In the case of the Atlantic margin of the eastern U.S. modeling studies using different seismic tomography models, different scalings of seismic velocity to density and different viscosity models of the mantle have resulted in significant discrepancies in estimates of the effects of dynamic topography on the evolution of this region. Some models[13-15] imply only dynamic topographic subsidence of the eastern seaboard of the U.S.; whereas in contrast, Moucha et al. [12], illustrated by the model discussed above, retrodict the dynamic topography evolution as a more complex pattern with regions of subsidence and uplift that vary spatially and temporally. As is made clear in Figure 1, as well as Figures 4 and 5, and well known previously ([17], among many others), the mid-Pliocene shoreline and shallow marine sediments are draped over a warped surface. This surface conforms well, at least in its northern part, with retrodictions of dynamic topographic uplift, not subsidence, of this region in the past 3 million years. The correspondence of these retrodictions with what is observed geologically thus provides a compelling and unambiguous test of the Moucha et al.[12] calculations of dynamic topography change.

Models that retrodict only dynamic topographic subsidence of the eastern seaboard[13-15] over this interval are not compatible with observed geology. This suggests that one or more of the assumptions employed in these models are incorrect. Furthermore, these models [14,15] integrate mantle flow over long timescales (>50 Ma) starting from the present, and thus any misfit with more recent times (<5 Ma), implies that these models are unlikely to retrodict dynamic topography correctly at any older time. This has significant implications for attempts to rectify the results of backstripping analyses of the New Jersey margin [3,10] with various other, nominally more global estimates of sea level derived from estimates of ridge volume changes [50], coastal onlap [51], ocean basin reconstruction [14], among other approaches [52,53], that infer higher long term global sea levels in the past (see summary by [14]). Our analyses do not support large amplitudes of dynamic topographic subsidence along the Atlantic shelf margin of North America, either on short time scales considered here or on longer time scale (30 Ma) investigated by Moucha et al.[12] and therefore do not support such rectifications.

The retrodictions from our mantle dynamics simulations imply that the east coast of the U.S. has been significantly impacted by dynamic topography change. It is important to note that the rates of change, as for example shown on Fig. 1, are not sustained rates over time, i.e. one cannot multiply the change in dynamic topography from 0 Ma to 3 Ma over some other interval to compute a total change in dynamic topography. Thus, the rate from 0 to 1 Ma is not one third that from 0 to 3 Ma; rather rates need to be computed per interval per location, reflecting the spatial and temporal evolution.

In the area of the Norfolk Arch where the largest amplitude of retrodicted dynamic topography change is centered, the rate of change per million years is about 60 m/Ma. This dynamic topography rate of change is about 3 times the maximum rate of change of the long term global sea level curve[50] since the base of the Jurassic, and is greater than about 85% of the rates of change of their short term global sea level curve based on a sampling of their curve at an interval of 0.1Ma. Thus the significant regression from the Albemarle and southern Salisbury Embayments since mid-Pliocene, which from the local sequence stratigraphic perspective would be directly linked to a significant global sea level fall, is modeled instead to be completely dominated by dynamic topographic uplift with little or no change in global sea level (Figure 5).

Similarly, the mid-Pliocene stratigraphy of New Jersey [45] is dominated not by transgressive sequences, as in the region from at least Virginia southwards, but by denudation and incision of earlier Miocene flooding surfaces and deposition of the Pensauken fluvial clastics. Clearly if New Jersey is rising out of the water at the same time that Virginia and points south are being flooded then some other processes, including, but perhaps not limited to, dynamic topography are controlling the sequence stratigraphy of this margin. These stratigraphic sequences, looked at locally, appear to be completely interpretable in terms of coastal onlap and offlap modulated by global sea level, but when examined on a more regional scale demonstrate that global sea level is not the only, and perhaps not even the dominant contributor to this pattern. To put this in perspective, the average slope of the shelf surface is about 0.05°±0.025°, based on the present width of the Atlantic shelf, measured from the shoreline to the -100m depth contour, even small changes (~20 to 40m) in dynamic topography beneath the shelf would move the shoreline by 10’s of km, i.e. on the scale of the shifts of coastal onlap often seen in seismic sequence stratigraphy and interpreted in terms of global sea level variations [51,54]. The regression of the Pliocene and younger sequences from the Albemarle Embayment, appears to specifically reflect such an effect.

Moucha et al.[12] stated that there is no such thing as a stable reference frame against which to measure global sea level. Coastal Plain geologists, particularly working from Virginia and farther south have long interpreted coastal scarps, similar to the Orangeburg scarp, but seaward and at lower elevations as reflecting previous highstands of sea level, with the descending heights of these scarps reflecting a progression in age and correlated maximum height of sea level associated with each of scarp ([55,56], among many others). An alternative interpretation presents itself based on the present analysis. If the long-term maximum height of sea level were to remain approximately fixed over the past 5 Ma, but the coast were to emerge as a result of changing dynamic topography, then the shore would move seaward, resulting in the observed stairstep pattern of observed coastal scarps. This is not to argue that global sea level has not varied, but rather to point out that there is a tendency to assume the underlying continental basement is relatively fixed on short time scales and hence to focus on global sea level, when, in fact, both are constantly moving relative to each other resulting in a complex signal that is not readily disentangled.

Was the height of global sea level during the maximum coastal onlap of the mid-Pliocene 35±18 m[8] higher than present sea level? Does the reconstruction of the Orangeburg Scarp provide compelling evidence for a more unstable East Antarctic ice sheet (see [21]) than often assumed? Our analysis does not allow an accurate estimate, given all of the uncertainties in various parameters that contribute to the adjustment of the east coast topography, but it should be clear that the Orangeburg Scarp, even when its height is adjusted for superimposed deflections [8], can not be used as a dipstick-like estimator of global sea level at this time. In fact, our retrodicted paleogeography at 3 Ma (Figure 5) matches the known distribution of mid-Pliocene marine strata in the Albemarle and southern Salisbury Embayments quite well with no change in global sea level at that time relative today. However we mismatch the equivalent Duplin and Charlton marine sequences farther south in Georgia (Figure 5) which clearly indicates that there remain unresolved sources of topographic perturbation in our analysis. Any assessment of the height of mid-Pliocene global sea level and its implications for East Antarctic ice sheet stability must be predicated on a global analysis of available data, and not on local investigations, which leaves the answers to the above questions unanswerable definitively, but should raise significant doubts about their veracity.

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Supplementary Information

Dynamic Topography Change of the Eastern U. S. since 4 Ma: Implications for Sea Level and Stratigraphic Architecture of Passive Margins

3-D lithospheric flexure modelling

The 3-D flexural response to the offshore sediment load for the entire region from George's Bank off Nova Scotia to the entire Gulf of Mexico is computed in order to capture longer wavelength contributions from regions outside our main area of interest. The 3-D flexural response is calculated by expanding the sediment thickness into spherical harmonics up to harmonic degree and order 360 (corresponding to a resolution scale - half-wavelength - of about 0.5 degrees). A Lanczos-smoothing of this harmonic expansion was applied to mute the Gibbs oscillations ('ringing') associated with this truncated representation. The 3-D elastic-isostatic flexure calculation (in spherical harmonics) uses the flexure parameters from the 2-D study of Pazzaglia & Gardner (1994 – ref. 7) which are a 40-km thick elastic plate with a Young's modulus of 70 GPa. The calculation used a sediment density of 2000 kg/m^3, mantle/lithosphere density of 3200 kg/m^3, and water density of 1000 kg/m^3. The effective submarine 'load'-density sitting on the elastic plate is the difference between the sediments and water and hence is equal to 1000 kg/m^3.

The 3-D flexural response to erosional unloading follows the same procedure as above based on an estimate of erosion over the past 5 Ma over the entire eastern U.S. modified after ref 39, east of 89°W.

Importance of hot, buoyant mantle under the eastern US

The principal, outstanding feature of the dynamic topography retrodictions presented in the main text is the pattern of variable uplift along the east coast of the US (Fig. 1). The origin of this uplift can be directly traced to existence of hot, buoyant material in the shallow mantle under this region, as well as the 'far-field' contribution from hot mantle under Bermuda. The impact of this active, buoyant material on the upper-mantle convective flow field is shown below in Fig. S1, where we observe that the centres of upwelling mantle under the eastern margin of the US are directly correlated with (and contributing to) the pattern of recent, post-Pliocene uplift of the coastal plain shown in Fig. 1.

The dynamical importance of this buoyant material was first made evident in previous calculations (Forte et al. 2007 – ref. 41) that focused on the impact of Farallon subduction under the central US, where shallow-angle upwelling mantle under the east coast resembled 'corner flow' above the subducting slab. We emphasize, however, that this is not passive return flow, because we find a source of deep buoyancy under the Atlantic mantle adjacent to the US coastal margin that extends to the Bermuda swell (see Fig. 4a in Forte et al. 2007). This upper-mantle buoyancy is an important contributor to the westward-directed upper-mantle flow under the eastern US shown below in Fig. S1.

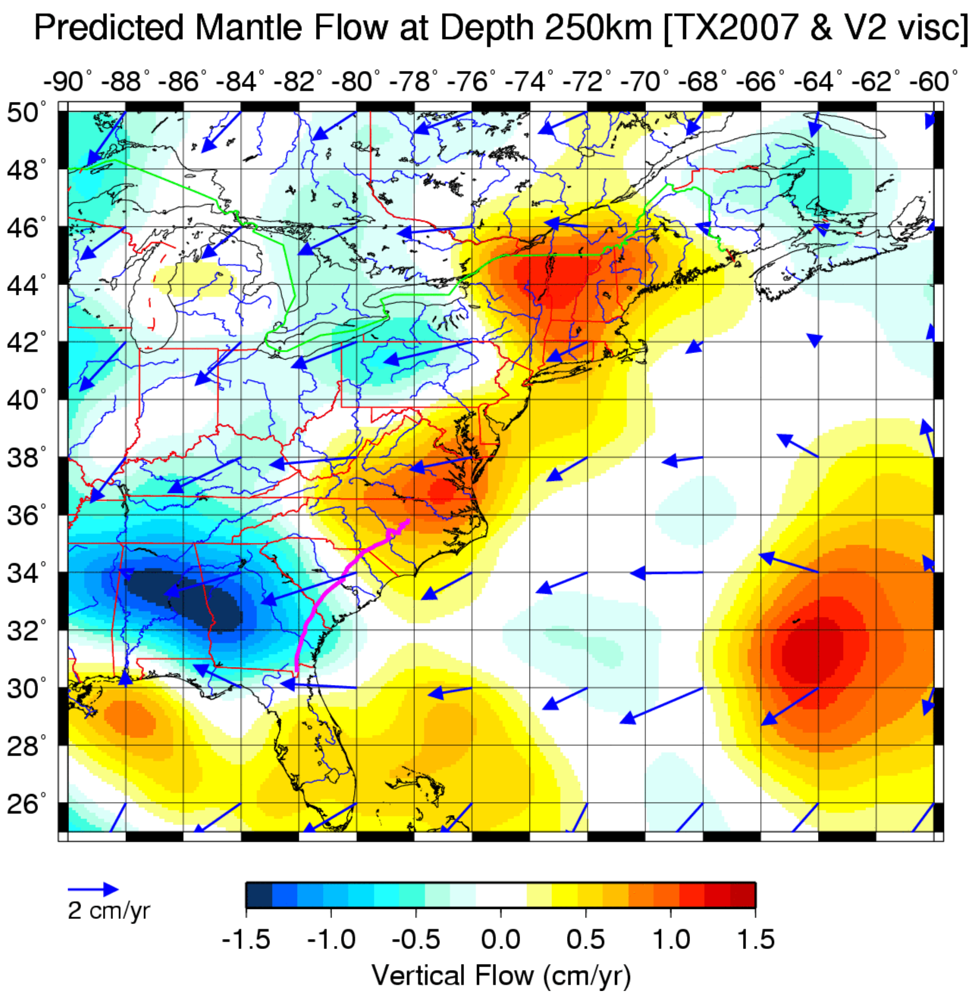

Figure S1. Present-day mantle convective flow in the asthenosphere under the east coast of the United States and adjacent Atlantic Ocean. The flow shown here at a depth of 250 km is predicted using the TX2007 joint seismic-geodynamic tomography model (Simmons et al. 2007 – ref. 43) and the V2 viscosity model (Forte et al. 2010 – ref. 42). The colour contours (scale bar at bottom centre) represent the vertical component of the flow and the blue arrows (scale at bottom left) the horizontal component. The mantle flow is represented in terms of a spherical harmonic expansion up to degree 128.

To further elucidate the dynamical importance of upper-mantle buoyancy for the evolution of east coast topography, we calculated the present-day dynamic topography due only to hotter-than-average material in the upper mantle under the Atlantic Ocean, shown below in Fig. S2. The mantle 'swell' due to buoyancy that is centred under Bermuda extends a considerable distance westward, towards the eastern margin of the US. This hot, buoyant material will be advected westwards (and upwards) by the prevailing mantle 'wind' (Fig. S1). As a consequence, there will be a progressive 'wave' of topographic uplift, associated with the westward transport of the topography signal in Fig. S2, that produces the variable post-Pliocene uplift shown in Fig. 1.

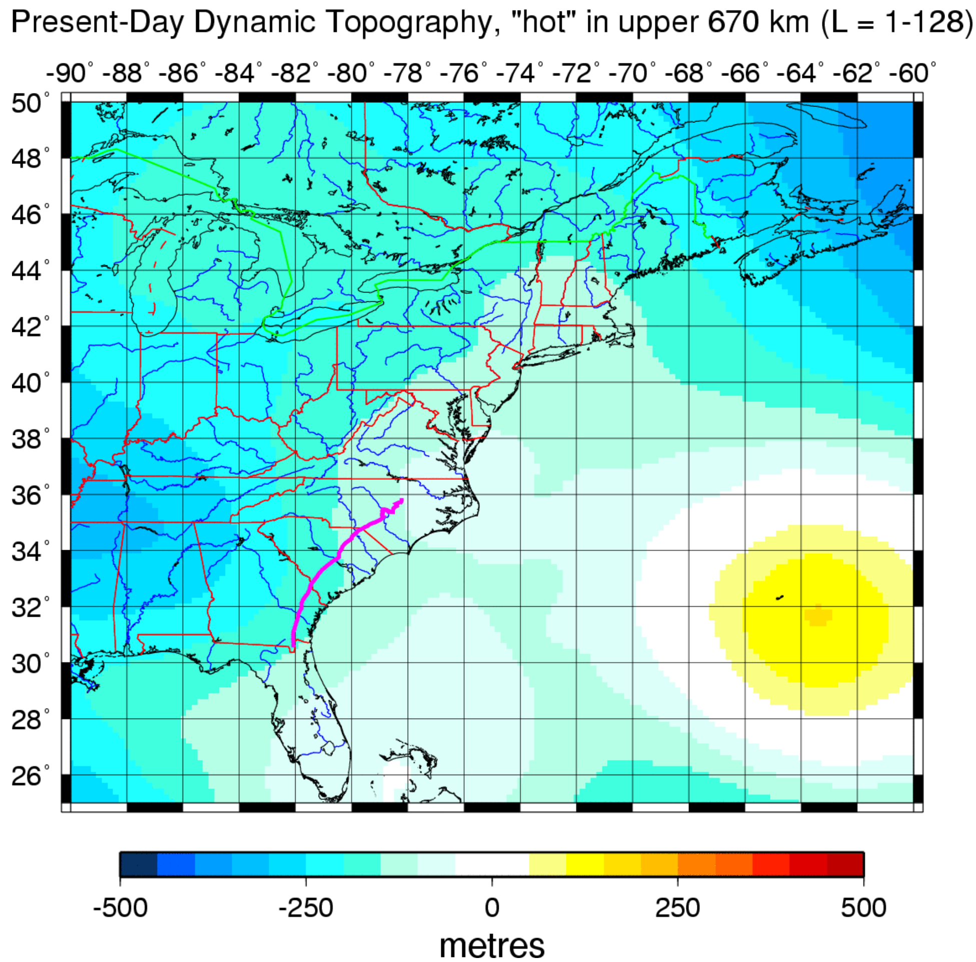

Figure S2. Present-day dynamic topography due only to hot, buoyant material in the upper mantle. This prediction is obtained in a mantle flow calculation in which we have stripping away all 'cold' heterogeneity (characterised by faster than average shear-wave velocity) in the joint tomography model TX2007 (Simmons et al. 2007 – ref. 43) and by using only the residual 'hot' heterogeneity in the upper mantle (down to 670 km depth). The V2 viscosity model (Forte et al. 2010 – ref. 42) is used in this calculation. The colour contours (scale bar at bottom) represent the vertical surface deflection. The mantle flow is represented in terms of a spherical harmonic expansion up to degree 128.